Recent progress in understanding of ice sheets, the Atlantic meridional overturning circulation, tropical forests and responses to ocean acidification

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Abstract

This article reviews recent scientific progress, relating to four major systems that could exhibit threshold behaviour: ice sheets, the atlantic meridional overturning circulation (AMOC), tropical forests and ecosystem responses to ocean acidification. The focus is on advances since the Intergovernmental Panel on Climate Change Fifth Assessment Report (IPCC AR5). The most significant developments in each component are identified by synthesizing input from multiple experts from each field. For ice sheets, key findings include: some degree of irreversible loss of part of the West Antarctic Ice Sheet (WAIS) may have already begun, but the rate and eventual magnitude of this irreversible loss is uncertain. For the early Holocene (sustained warming ~2K above pre-industrial, at the maximum warming level suggested by the Paris climate agreement), WAIS mass loss rates comparable to present-day have been inferred, but without a WAIS collapse. The observed AMOC overturning has decreased from 2004-2014, but it is unclear at this stage whether this is forced or is internal variability. New evidence from experimental and natural droughts has given greater confidence that tropical forests are adversely affected by drought. The ecological and socio-economic impacts of ocean acidification are expected to greatly increase over the range from today's annual value of around 400 up to 650 ppm CO_2 in the atmosphere, with rapid development of aragonite undersaturation at high latitudes. Tropical coral reefs are vulnerable to the interaction of ocean acidification and temperature rise, with uncertain survival at 2˚C warming above pre-industrial. Across the four systems studied, however, quantitative evidence for a difference in risk between 1.5˚C and 2˚C warming above pre-industrial levels is limited.

1. Introduction

While some aspects of climate change can be viewed as becoming proportionately larger with increasing forcings, other aspects may feature more complex, nonlinear behaviour (e.g. Lenton et al. 2008). This can include abrupt and/or irreversible change, which may be associated with key thresholds. Such behaviour must be considered differently in assessments of the potential benefits of mitigation: it implies, for example, that certain impacts could be significantly different above certain levels of anthropogenic interference.

Clear evidence of threshold behaviour in the earth system is seen in the paleoclimate record (e.g. McNeall et al. 2011). For example, central Greenland temperatures inferred from ice cores show relatively abrupt changes, even though the forcing over this period has evolved smoothly. These changes are thought in part to be associated with changes in the ocean's thermohaline circulation (Broecker 2003), although strong paleoclimate evidence for threshold behaviour in major ice sheets and methane reservoirs in particular also exists (McNeall et al., 2011).

This study focuses on four major systems that may feature threshold behaviour: ice sheets, the Atlantic Meridional Overturning Circulation (AMOC), tropical forests, and ecosystem responses to ocean acidifcation. The risk of significant change in these systems is not necessarily linked to large-scale warming alone. Patterns of precipitation can be important for the AMOC and tropical forests; tropical forests are also strongly affected by anthropogenic land-use and a direct effect of carbon dioxide, while ocean acidification arises directly from increased atmospheric CO_2 (although its impacts combine with those of ocean warming) and West Antarctic Ice Sheet (WAIS) stability is influenced by changes in ocean circulation.

Consequences of change in these systems range from amplified global warming through altered climate patterns, elevated sea-level and direct loss of biodiversity and ecosystem services (see individual sections below for details). These systems can in principle interact with each other (Lenton et al., 2008), although this is explored in only a few studies.

Here we report primarily on new literature subsequent to that presented in the IPCC Fifth Assessment Report, AR5. We also briefly consider (in the Conclusions) the difference in risk between 1.5K and 2K global mean warming above pre-industrial levels. This review was prepared using an iterative approach, by specialists both within and external to the Met Office Hadley Centre. Initial drafts of each section were prepared by the Met Office, then sent to external experts for review and editing (except that for Ocean Acidification; prepared by experts in the National Oceanography Centre in Southampton, and the Plymouth Marine Laboratory). The sections were revised accordingly by the Met Office, then sent to the external experts for a second review.

Each system is addressed in a separate section below, each with the following subsections: Introduction (the key issues for that system); Observations (relevant real-world observations); Potential for significant change (literature addressing the question of how likely substantial change is); Consequences (of significant change); Cautions (key scientific uncertainties); and Comparison with AR5. The key conclusions are summarised in Table 1.

2. Ice Sheets

a Introduction

Ice-sheet mass loss from the Greenland and Antarctic ice sheets, is of concern due to its impact on global sea level (Alley et al. 2005), and potential amplification of global warming over long timescales as low-albedo land surface is exposed (Hansen et al. 2008). The ice sheets are the largest potential source of future sea level rise on many time-scales as well as the most uncertain. They contain enough ice to raise mean sea level by some 65 m. In addition, rapid mass loss may have an influence on ocean circulation.

In a state of equilibrium, an ice sheet loses mass through the melting and calving of its outlet glaciers at the same rate as it gains mass through the accumulation of snowfall (Alley et al, 2005). Increased ice sheet mass loss occurs through two main mechanisms. Increased surface melt is largely driven by higher air temperatures and currently only affects the Greenland ice sheet mass balance. ‘Dynamic thinning’ (i.e. losses due to increased solid ice discharge into the ocean) involves glacier acceleration and consequent increases in iceberg calving for marine-terminating glaciers. This may be induced by increased surface melt but also by ocean-related processes such as changes in ocean heat content and/or circulation beneath floating ice or at the terminus (Gille 2014).

Ice shelves, the floating portions of outlet glaciers, play a key role in modulating the mass balance of the Antarctic ice sheet. They buttress the inland glaciers, helping to control the rate of ice leaving the continent and entering the ocean (Dupont; Alley 2005). Ice shelves are exposed to the underlying ocean and may weaken as ocean temperatures rise. If they melt rapidly or break away, ice flow accelerates, causing net ice-sheet mass loss (De Angelis; Skvarca 2003). Much of the West Antarctic Ice Sheet (WAIS) is grounded on bedrock below sea level on retrograde slopes (deeper inland). This configuration is believed to be inherently unstable and sensitive to small changes in the grounding line (where the ice begins to float; Mercer 1968; Schoof 2007). This is known as the marine ice sheet instability hypothesis.

The Greenland ice sheet is a “relict” from the last Glacial that ended about 12 K years ago. The altitude of the ice sheet interior maintains the persistently cold temperatures required for the ice sheet to survive. There is a temperature threshold above which the Greenland ice sheet is no longer viable (Gregory; Huybrechts 2006; Robinson et al. 2012). This is because, as temperatures increase, so does melting, which results in a lowering in surface elevation, causing further warming (atmospheric temperature decreases with altitude). This positive feedback is known as the small ice cap instability (Crowley; North 1988).

Key issues addressed by recent studies include: what the observed ice-sheet loss implies for the rate of future sea-level change, the potential long-term sea-level rise, and the possibility of abrupt or irreversible changes.

b Observed recent changes

Currently, ice loss from the Amundsen Sea sector of the West Antarctic Ice Sheet (WAIS) contributes 0.28 mm yr^{-1} to global sea-level rise. Pine Island glacier (Favier et al. 2014) and

Thwaites glacier (Joughin et al. 2014) are the principal outlets of the WAIS that have rapidly thinned, retreated, and accelerated. The spatial pattern of ice shelf thinning in the Amundsen Sea, suggests that the loss of grounded ice is the direct result of increased basal melting of the ice shelf, as a consequence of the inflow of warm water from the southern Pacific (Ha et al. 2014; Jacobs et al. 2011). Multi-decadal warming at the seabed in the Bellingshausen and Amundsen seas is linked to increased heat content and to a shoaling of the mid-depth temperature maximum over the continental slope, allowing warmer, saltier water greater access to the continental shelf in recent years (Schmidtko et al. 2014). Since about 2009, the Southern Antarctic Peninsula has been contributing significantly to sea level rise at a near-constant rate of 0.16 mm yr^{-1} (Wouters et al. 2015). The onset of this sudden and rapid mass loss appears to have a similar origin to that seen in the Amundsen Sea sector. The bedrock configuration is such that the mass loss is likely to be sustained for years to decades into the future, for this sector of Antarctica.

In addition there have been synchronous advances and retreats of the tide-water glaciers of the East Antarctic Ice Sheet (EAIS) (Miles et al. 2013), associated with changes in the Southern Annular Mode (SAM).

The Greenland ice sheet (GrIS) is losing mass as a result of both increased runoff due to surface melting and increased ice discharge from marine-terminating outlet glaciers (Rignot et al. 2008; Rignot et al. 2011; Sasgen et al. 2012; van den Broeke et al. 2009). The Greenland mass loss over the period 2000-2012 contributed about 0.7 mm yr^{-1} of sea level rise. The rate, has however, been accelerating over this time period such that for 2009-2012, it was 1.05 mm yr^{-1} (Enderlin et al. 2014). The relative contribution of ice discharge (dynamic thinning) to total loss decreased from 58% before 2005 to 32% between 2009 and

2012. As such, 84% of the increase in mass loss after 2009 was due to increased surface runoff as opposed to increased discharge (Enderlin et al, 2014). These observations support recent model projections that changes in surface mass balance driven, primarily, by increases in air temperature, rather than ice dynamics, will likely dominate the ice sheet's contribution to 21st century sea level rise (e.g. Goelzer et al. 2013; Vizcaino et al. 2015).

The glaciers in the southeast and northwest of Greenland sped up between 2000 and 2005 and have since stabilised or slowed (Enderlin et al, 2014). The slow down in the southeast has been compensated for by the northeast Greenland ice stream, which extends more than 600 km into the interior of the ice sheet, and is now undergoing sustained dynamic thinning, linked to regional warming, after more than a quarter of a century of stability (Khan et al. 2014). This sector of the Greenland ice sheet is of particular interest, because the drainage basin area covers 16% of the ice sheet and numerical model predictions suggest no significant mass loss for this sector, leading to a possible under-estimation of future global sea-level rise (Khan et al, 2014). As for the Southern Antarctic Peninsula, the geometry of the bedrock and monotonic trend in glacier speed-up and mass loss suggests that dynamic loss of grounded ice in this region will continue in the near future (Khan et al., 2014).

c Potential for significant change

The ice sheets can lead to dangerous climate change through accelerated discharge of freshwater to the ocean, and their potential irreversibility. Accelerated discharge, particularly from a possible marine ice sheet instability, has implications for the predictability of future sea level rise. New studies here have focussed on the WAIS.

An ice flow mode l (Favier et al. 2012) reveals that Pine Island Glacier's grounding line is probably engaged in an unstable 40 km retreat (Figure 1; Favier et al., 2014). The associated mass loss increases substantially over the course of the simulations from an average value of 0.05 mm yr^{-1} observed for the 1992-2011 period, up to and above 0.28 mm yr^{-1}, equivalent to 3.5-10 mm mean sea-level rise over the next 20 years (Favier et al., 2014). They find that mass loss remains elevated from then on, ranging from 0.16 to 0.33 mm yr^{-1}. New paleoclimate evidence (Johnson et al. 2014) for the early Holocene (a period of sustained warming about 2K above pre-industrial) has revealed mass loss from the Pine Island Glacier at a rate comparable to present-day loss, but no collapse. Simulations for the adjacent Thwaites glacier, also in the Amundsen Sea embayment, indicate future mass losses are moderate <0.25 mm yr^{-1} over the 21st century but generally increase thereafter (Joughin et al. 2014). Paleoclimate evidence has been used to provide an empirical assessment timescale of collapse of the Pine Island and Thwaites Glacier catchments in West Antarctica (Kleman; Applegate 2014). The study suggests that the Pine Island Glacier may experience a minor collapse over its main trunk, but the bed topography favours a less dramatic retreat thereafter. On the other hand the Thwaites Glacier is probably not as close to a threshold as Pine Island Glacier, but once efficient drainage has progressed inwards a full collapse of the area may occur. The likely time scale for collapse is the time required for 100-200 km of grounding line retreat in the Thwaites Glacier system plus 100-300 years for an actual collapse event (Joughin et al 2014). Except possibly for the lowest-melt scenario used in the simulations, the results indicate that early-stage irreversible collapse has already begun (Joughin et al 2014).. Less certain is the time scale, with the onset of rapid (>1 mm yr^{-1} of sea-level rise) collapse in the different simulations within the range of 200 to 900 years**.**

For Antarctica as a whole, there is new evidence (Weber et al. 2014) for periods of relatively abrupt Antarctic mass loss following the Last Glacial Maximum, possibly associated with a positive feedback involving ocean heat transport. Further back in time, during the Miocene, when CO2 levels fluctuated between 280 and 500 ppm (equivalent to pre-industrial and a value that will be reached in the next few decades) there is evidence that the Antarctic ice sheet, including East Antarctica, experienced a volume loss on the order of 20 m of sea level equivalent compared to present-day (Levy et al. 2016).

d Potential consequences

The eventual (partial) collapse of the WAIS seems likely, leading to a global sea level rise of up to 3.3 m (Bamber et al. 2009) with an additional contribution from parts of East Antarctica that will be affected by WAIS drawdown. This inference is supported by several lines of recent observational evidence. It has been suggested that a critical threshold for grounding line retreat has already been passed for glaciers in the Amundsen Sea sector (Rignot et al. 2014). High ice shelf thinning rates for this and the Bellinghausen Sea sector of West Antarctica over the last two decades (Paolo et al. 2015) combined with the dramatic shift in mass imbalance of the Southern Antarctic Peninsula (Wouters et al, 2015) also point to a widespread shift in behaviour for this region.

Surface melt from the Greeland Ice Sheet may influence local ocean circulation and consequently local sea level change, perhaps by 5 cm, in the North-West Atlantic (Howard et al. 2014; Swingedouw et al. 2013).

Substantial topographic change associated with melt of the Greenland ice sheet can be expected on the 200-500 years timescale (Ridley et al. 2005; Vizcaino et al. 2008). The reduction in the topographic barrier to westerly flow will lead to a wider impact on the Northern Hemisphere climate. In addition the local albedo change will cause vegetation changes which will have a feedback on the rate of deglaciation (Stone; Lunt 2013).

On centennial to millennial time scales Antarctic Ice Sheet melt can moderate warming in the Southern Hemisphere, by up to 10ºC regionally, in a 4 x CO_2 scenario (Swingedouw et al. 2008). This behaviour stems from the formation of a cold halocline in the Southern Ocean, which limits sea-ice cover retreat under global warming and increases surface albedo, reducing local surface warming. In addition, Antarctic ice sheet melt, by decreasing Antarctic Bottom Water formation, restrains the weakening of the Atlantic meridional overturning circulation, which is an effect of the bi-polar oceanic seesaw (Pedro et al. 2011). Consequently, it appears that Antarctic ice sheet melting strongly interacts with climate and ocean circulation globally. It is therefore necessary to account for this coupling in future climate and sea-level rise scenarios.

e Cautions (uncertainties).

While substantial progress in understanding has been made, it is still unclear what the recent observed changes imply for long-term future ice-sheet loss (Nick et al. 2009). New observations suggest that there may be a natural cycle of increase and decrease in the rates of mass loss from coastal glaciers (Murray et al. 2010), so short-term trends should not necessarily be extrapolated into the future (Wouters et al. 2013). Indeed many Greenland glaciers, which accelerated in the early 2000s have since slowed (Enderlin et al. 2014; Moon et al. 2012). There is a potential that solid earth movement, in response to ice loss, may influence the bedrock slopes and reduce further ice loss from the West Antarctic Ice Sheet (Konrad et al. 2015).

f Comparison with AR5 (90)

Of the key findings summarised in Table 1, the main new points since AR5 are: observational evidence (Enderlin et al., 2014) that, from Greenland, the proportion of loss from surface melt has increased, becoming more consistent with long term model projections; evidence that some degree of irreversible loss from the WAIS may have begun (Favier et al., 2014, Joughin et al, 2014, Rignot et al, 2014, Wouters et al 2015); and indications that the East Antarctic Ice Sheet) (Miles et al., 2013) and that northeast Greenland (Khan et al., 2014) may be more sensitive to climate change than previously expected.

2. AMOC

a Introduction

The Atlantic Meridional Overturning Circulation (AMOC) transports large amounts of heat northwards in the Atlantic Ocean, resulting in a milder climate in northwest Europe and the North Atlantic than would otherwise be experienced (for recent reviews of AMOC behaviour and observations see Srokosz et al. 2012; Srokosz; Bryden 2015). The IPCC AR5 report concludes that it is very likely that the AMOC will weaken over the 21st century, although there is a large spread among climate models in the predicted weakening. A large or rapid reduction in the AMOC would likely have substantial impacts on global climate, although a collapse of the AMOC by 2100, however, was judged as very unlikely (Collins et al. 2013). These top-level assessments have not changed since the previous IPCC assessment.

b Observed recent changes

The RAPID-MOCHA array has been observing the AMOC at 26°N since 2004 and now has acquired a decade of data (McCarthy et al. 2015b; Rayner et al. 2011). This dataset has revealed large variability on timescales from daily to interannual (see Figure 2). This included a large (30%), temporary decrease in AMOC strength over 2009-2010 (Bryden et al. 2014; McCarthy et al. 2012), which resulted in cooling in the upper North Atlantic Ocean in 2010 north of the latitude of the RAPID array and warming to the south (Bryden et al. 2014; Cunningham et al. 2013). This decrease began with a strengthening of the upper mid-ocean recirculation in early 2009 and was compounded by a slowdown in the northward Ekman transport and Gulf Stream flow in late 2009 and early 2010 (accounting for 61%, 27% and 12% of the slowdown, respectively; Bryden et al. 2014). This decrease was well outside the range predicted for interannual AMOC variability in coupled ocean-atmosphere models (McCarthy et al. 2012; Roberts et al. 2014). Roberts et al. (2013) reproduced this AMOC decrease using an initial condition ensemble of ocean simulations driven by observed surface forcing (albeit with too weak an AMOC), suggesting that the atmosphere may have had a dominant role in the temporary AMOC decrease. However the origin of, and complete explanation for, the 2009-10 event remain uncertain. To-date no explanations have fully accounted for the changes in Lower North Atlantic Deep Water (LNADW at 3000 to 5000m depth) and the lack of change in the Upper North Atlantic Deep Water (UNADW between 1000 and 3000m depth) described by McCarthy et al. (2012).

The links between changes in the AMOC, upper ocean heat content and atmospheric response represent an active area of research. For example, the ocean has been implicated in the re-emergence of sea surface temperature anomalies from the winter of 2009-10 during the following early winter season of 2010-11, which contributed to the persistence of the negative winter North Atlantic Oscillation (NAO) and wintry conditions in northern Europe (Taws et al. 2011). Such behaviour may lead to improved predictions of the NAO and winter conditions (Maidens et al. 2013; Scaife et al. 2014).

The AMOC overturning has also decreased from 2004-2014 (Figure 2; Srokosz; Bryden 2015); the majority of this was due to a weakening of the geostrophic flow (Smeed et al. 2014; who analysed the first eight and a half years of data). This trend has been associated with decreases in subsurface density in the subpolar gyre, similar to those seen in climate models when there is a reduction in the AMOC (Robson et al. 2014). It is unclear at this stage whether the decrease is forced. Statistical tests on the observations (Smeed et al., 2014) suggested that the AMOC decrease is significant, even if the low AMOC event of 2009-10 is excluded. Roberts et al (2014) found similar trends as part of natural variability in 2 out of 14 global climate models, and in all models considered when corrections are made to include more realistic high frequency variability. They concluded that more than a decade of observations are required to detect and attribute an anthropogenic weakening of the same trend as observed. Send et al. (2011) observed a decreasing trend in the transport of the deep western boundary current at 16^{o}N (one component of the AMOC) over a similar period.

A recent paper (Rahmstorf et al. 2015) suggested that this trend is part of an 'exceptional slowdown' of the AMOC. They find a relationship between sea surface temperatures and the AMOC in a climate model and then use reconstructions of surface temperatures from paleoclimate records to suggest that there has been a weakening that is unprecedented over the last 1000 years. There are, however, inherent uncertainties around both the relationship used and the temperature reconstructions. Ultimately, all proxies for the AMOC, such as temperature or coastal sea level (Ezer 2015; Frajka-Williams 2015; McCarthy et al. 2015b) need to be tested and verified against direct observations, over the time scales of interest, if they are to be used to infer its behaviour over longer periods.

Future observations and research will improve our assessments of past and on-going AMOC changes. In this context note that the Overturning in the Subpolar North Atlantic Program (OSNAP; see http://www.o-snap.org/ and http://www.ukosnap.org/) deployed instruments in 2014 along a line from Canada to Greenland to Scotland, to observe the AMOC in the subpolar gyre, complementing the 26.5˚N observations in the subtropical gyre. Meanwhile in the South Atlantic there are trans-basin observations of the AMOC beginning to be made at 34.5˚S (SAMBA – South Atlantic MOC Basin-wide Array; Ansorge et al. 2014; Meinen et al. 2013). Recently, a new component of the AMOC, the so-called East Greenland spill jet, has been identified from a year of mooring observations (von Appen et al. 2014), but its importance in the long-term for the overall AMOC remains to be confirmed.

c Potential for significant change

Paleoclimate studies have suggested that abrupt changes to climate may have been caused by the AMOC switching from an “on” state, where it transports heat northwards in the Atlantic, to an “off” state (Rahmstorf 2002).. It is thought that these abrupt changes may be related to the existence of bistability (where both “on” and “off” states of the AMOC can exist for a given forcing) as predicted by theoretical models of the Atlantic (e.g. Stommel 1961), Earth system models of intermediate complexity (Rahmstorf et al. 2005) and one study with a coarse resolution global circulation model (Hawkins et al. 2011).

There have been many studies suggesting that the stability of the AMOC might be affected, or even controlled, by whether the AMOC imports or exports fresh water from the Atlantic, since this can indicate the presence of a positive or negative advective feedback. De Vries and Weber (2005) found that the fresh water transport by the AMOC into the Atlantic (Fov) was an important indicator of stability in their experiments, however the precise role it plays is still unclear Other factors have subsequently been found to be important in determining AMOC stability. For example, Jackson (2013) found that, while Fov does partially indicate the sign of the advective feedback in a GCM, the transport of fresh water by the gyres can also play a crucial role. Swingedouw et al. (2013) also found that gyre transports can affect the magnitude of AMOC reduction. The presence of eddies in an eddy resolving model can also affect the response of the fresh water transport (den Toom et al. 2014).

d Potential consequences

A collapse in the AMOC would cause a large relative cooling over the North Atlantic, which would have wide-ranging impacts. As well as previously documented impacts on the physical system, such as cooling in the northern hemisphere and a shift in the Inter tropical Convergence Zone which cause substantial changes in tropical precipitation, (Jackson et al. 2015; Vellinga; Wood 2008), there have been a couple of recent studies detailing the impacts on the carbon cycle and vegetation. Bozbiyik et al (2011) showed that a reduction in the AMOC has a large impact on the Amazon with reduced precipitation causing large reductions in vegetation. Parsons et al. (2014), on the other hand, found that a reduction in the AMOC caused an increase in vegetation over the Amazon, due to a change in precipitation seasonality (despite a reduction in annual mean precipitation).

Other studies have concentrated on impacts over Europe. Woollings et al. (2012) showed that models with a greater reduction in AMOC strength due to increased greenhouse gases have greater increases in strength of the North Atlantic storm track, implying an increase in the number of winter storms across Europe. This was also found in the study by Jackson et al (2015), who also showed that the increase in winter storms resulted in greater precipitation over western coasts in Northern Europe, despite a general reduction of precipitation over the northern hemisphere from a cooling-induced reduction in evaporation. They also found regional changes in summer precipitation across Europe, similar to those associated with

Atlantic sea temperature found by Sutton and Dong (2012). Haarsma et al. (2015) examined the relationship between European atmospheric circulation and the AMOC across the CMIP5 ensemble. They also found an influence of AMOC strength on European summer precipitation and cloud cover.

One impact of the AMOC suggested recently is its role in the so-called global warming "hiatus" (Chen; Tung 2014), though various other oceanographic explanations for the hiatus have been proposed. A recently observed impact is the reduction in uptake of CO_2 by the Atlantic Ocean due to the weakening of the AMOC over the period 1990 to 2006 (Perez et al. 2013), which has potential consequences for the rate of global warming.

Another recent focus of attention has been the role of the AMOC in sea level rise (SLR) on the eastern seaboard of the USA (Ezer 2015; Goddard et al. 2015; McCarthy et al. 2015a). In particular, Goddard et al. (2015) demonstrate that the 2009-10 downturn in the AMOC led to an unprecedented 12.8 cm sea level rise along the coast north of New York over the same period. They show that this rise was a 1-in-850 year event. Furthermore, they note that, "Unlike storm surge, this event caused persistent and widespread coastal flooding even without apparent weather processes. In terms of beach erosion, the impact of the 2009–2010 SLR event is almost as significant as some hurricane events."

e Cautions

Several studies have shown that many GCMs have biases in the fresh water transport of the AMOC (importing instead of exporting fresh water), and that this might affect the simulated stability of the AMOC. The source of this bias is unclear. Jackson (2013) attributed the bias to an over-evaporative Atlantic in the model and notes the difference from observations in salinity profiles in the South Atlantic. Liu et al. (2014) suggested that the presence of a double Atlantic ITCZ (a common GCM bias) results in a tropical salinity bias that stabilises the AMOC. Another source of uncertainty is the transport of saline water from the Indian Ocean to the Atlantic by eddies that are shed from the Agulhas current. Current GCMs do not resolve the scales required to correctly represent these eddies, but a recent study by Biastoch and Böning (2013) used a high resolution nested model to resolve this region. They found that a southwards shift of the southern hemisphere westerlies (as is expected to occur under anthropogenic climate change) results in a decrease in salinity transport into the Atlantic, however this change in salinity is small and has little impact on the AMOC. The lack of eddy-resolving resolutions in current GCMs might also have an impact on the transient response of the AMOC to increased freshwater input (Weijer et al. 2012).

There is also substantial uncertainty about the future inputs of freshwater into the Atlantic, particularly since the climate models lack dynamic ice sheet models which could substantially speed up the input of freshwater from the Greenland ice sheet. Separate studies including additional freshwater inputs from the Greenland ice sheet find that projected changes do not have major impacts on the AMOC, although there is uncertainty about future changes in freshwater fluxes from Greenland (Bamber et al. 2012). A study as part of the European project FP7 THOR found that the MOC became less sensitive to fresh water inputs when CO_2 levels were high, because of increases in stratification caused by warming and changes in the wind-driven circulation (Swingedouw et al. 2015). Another recent study suggests that increased precipitation over the Arctic, leading to increased freshwater flux into the North Atlantic could also affect the AMOC (Bintanja; Selten 2014; see Methods).

f Comparison with AR5

The main development since the publication of AR5 has been the updated observations of overturning from the RAPID-MOCHA array (Smeed et al. 2014; Srokosz; Bryden 2015), which shows a decline over the period 2004-2014 (although it is unclear at this stage whether or not this is forced). The other key finding is the unprecedented rise in US east coast sea level associated with the 2009-10 downturn in the AMOC.

3. Tropical forests

a. Introduction

Tropical forests regulate and supply to society a range of services, which bring benefits at global to local scales. As well as sustaining high biodiversity they influence climate through biogeochemical (carbon cycle) and biophysical (water and energy) mechanisms. Over the period 1990-2007, tropical intact forests took up carbon at the rate of 1.2 ± 0.4 Pg C year^{-1}, (corresponding to about half the global carbon sink), compared with 0.50 ± 0.08 Pg C year^{-1} by the boreal forests (Pan et al. 2011), and around 1.1 ± 0.8 Pg C year^{-1} losses of forest carbon stocks to the atmosphere through land use change over 2000-2009 (Settele et al. 2014). However, large droughts can cause elevated mortality rates, especially for larger trees (da Costa et al. 2010; McDowell; Allen 2015; Nepstad et al. 2007; Phillips et al. 2010) and temporary shifts from ecosystem carbon sink to carbon source (Gatti et al. 2014; Lewis et al. 2011; Phillips et al. 2010). Estimates of the impact of the 2005 and 2010 droughts (mostly through increases in tree mortality during and lagging the droughts) stand at 1.6 and 2.2 Pg C, respectively (Lewis et al. 2011; Phillips et al. 2009).

Tropical forests are subject to interacting effects from atmospheric CO_2, climate and land-use change (e.g. Coe et al. 2013). Land-use change effects include direct deforestation, and accidental 'leakage' fires. Forest fragmentation (an important by-product of deforestation) lengthens the forest edge, accelerating the rate of forest erosion by fire. Deforestation increases albedo and reduces evapotranspiration, altering climate both locally and downwind; aerosols from deforestation fires may also reduce rainfall (Marengo et al. 2011). Climate change could alter vegetation productivity and mortality, both directly, and indirectly by modifying fire behaviour. Increased atmospheric CO_2 may increase tree growth (where nutrients are not limiting), but also increase tree mortality from lianas (vines). The full vegetation response to CO_2 and climate changes may take decades to be completely realised (Jones et al. 2009).

The AR5 finds that large-scale dieback due to climate change alone is unlikely by the end of this century (*medium confidence*). However, it states with *medium confidence* that "severe drought episodes, land use, and fire interact synergistically to drive the transition of mature Amazon forests to low-biomass, low-statured fire-adapted woody vegetation" (Settele et al. 2014). New research has largely, but not exclusively, focused on the Amazon: due in part to early climate model projections of climate-driven Amazon dieback (Cox et al. 2000). Severe Amazonian droughts in the last decade have provided insights on forest responses to extreme dry conditions. In addition to forest and climate monitoring, throughfall exclusion and prescribed-burn experiments have allowed in-situ study of the effects of longer-term drought and fire. Numerical studies have also increased in number and progress has been made in putting the early results into context.

b. Observations

New studies have given greater confidence that the Amazon represents a long-term net carbon sink (Brienen et al. 2015; Espirito-Santo et al. 2015; Gatti et al. 2014), but also suggest that its strength has weakened progressively as tree mortality rates increase (Figure

3).

The response of trees to elevated CO_2 remains uncertain. Some recent longer-term studies of tropical tree rings (Battipaglia et al. 2015; Groenendijk et al. 2015; van der Sleen et al. 2015) have found no evidence for sustained increases in tree growth or carbon uptake, but as Brienen et al. (2012) point out, tree-ring studies are subject to biases which preclude robust statements about ecosystem-level changes. So far, multi-decadal plot data been used systematically to probe recent growth trends at continental scale only in Amazonia (Brienen et al. 2015). Here they indicate a long-term increase in growth rates since the 1980's, as well as a lagging increase in mortality rates, consistent with a long-term growth stimulation, such as by CO_2.

There is greater confidence that Amazon forests are adversely affected by drought. There has been new work on the response to the 2010 drought, and also the 1997 and 2005 events (Tomasella et al. 2013). A new attribution study (Shiogama et al. 2013) of the 2010 drought showed that, while sea surface temperature anomalies in the tropical Pacific and Atlantic likely increased the probability of drought (in addition to biomass burning; Marengo et al. 2011), unforced atmospheric variability probably also played a large role. Atmospheric measurements (Gatti et al. 2014) confirmed earlier plot-based findings (Lewis et al. 2011; Phillips et al. 2010) that the Amazon switched from a temporarily from a net carbon sink to a source during the 2010 drought. Compared to these short-term natural droughts, the impact was seen to be much stronger in the long-term persistent experimental droughts induced by a forest throughfall exclusion experiment in eastern Amazonia (da Costa et al. 2014), and by 2014, 13 years of 50% throughfall exclusion at Caxiuana had caused a cumulative biomass loss of 45.0 ± 2.7% (Rowland et al. 2015). Consistent with previous suggestions that effects of a single drought persist for several years (Phillips et al. 2010; Saatchi et al. 2013), even during the anomalously wet year of 2011, the Amazon was still estimated to be carbon neutral overall (Gatti et al. 2014; possibly due to lagged effects of the 2010 drought).

Drought mortality, especially in larger trees, is a major pathway for carbon release (da Costa et al. 2010; Nepstad et al. 2007; Phillips et al. 2010), but underlying mechanisms are not well understood (Meir et al. 2015), and poorly represented in current vegetation models (Powell et al. 2013). However, hydraulic failure is suggested as the primary cause from the Caxiuana drought experiment (Rowland et al. 2015). A study of detailed plot-level responses to the 2010 drought in several sites, compared to other years (Doughty et al. 2015) suggested that trees may prioritise growth in response to reduced photosynthesis from short-term drought, leaving some trees more vulnerable to mortality. In contrast, the long-term (> 12 years) response to persistent experimental rainfall exclusion (Rowland et al. 2015), shows no decline in photosynthetic capacity (although photosynthesis may have declined if mean stomatal conductance declined), but an increase in leaf dark respiration in tree taxa vulnerable to drought mortality (possibly a sign of drought stress). It has been suggested that early warning of drought mortality events may be plausible based on observations of tree properties (Camarero et al. 2015). Overall, the AR5 viewpoint of persistent drought causing a shift towards lower statured, low-biomass forest is retained (Rowland et al. 2015).

Drought can also cause abrupt increases in fire-induced tree mortality (Brando et al. 2014). More than 85,500 km^2 of the southern Amazon was burnt by understorey fires during 1999-2010, with evidence for a strong climate control on fire (Morton et al. 2013).

Forest responses to warming remain uncertain, and more forest warming field experiments are needed (Cavaleri et al. 2015). In one such experiment (Slot et al. 2014), although respiration increased with warming, thermal acclimation did occur. A new meta study integrating experimental and observational results (Vanderwel et al. 2015) suggests that acclimation could potentially half increases in leaf dark respiration over the century, compared with null model expectations that ignore acclimation. On the other hand, a global-scale analysis of interannual variability has suggested (Anderegg et al. 2015) that nighttime respiration in tropical forests may be highly sensitive to warming.

Some new observational studies have found substantial reductions in evapotranspiration in some (da Silva et al. 2015; Oliveira et al. 2014; Panday et al. 2015), but not all (Rodriguez et al. 2010) deforested regions. The full effects of deforestation over the Xingu river basin may have been masked by climate variability (Panday et al. 2015).

c. Potential for significant change

A recent review of wider sources of evidence (Coe et al. 2013) identified South/South-East Amazonia as particularly vulnerable: due to high deforestation rates locally and in the upwind savanna region; its susceptibility to small climate shifts (being in a transitional climate zone between forest and savanna); and greater climate model agreemement on future drying in this region (compared to the West Amazon). A review for African rainforests (Malhi et al. 2013) highlighted similar points: while deforestation rates have historically been relatively low over Africa, there is potential for significant future increases; and African forests have climate close to the limit of rainforest sustainability. Models tend to predict drying(wetting) over western(central) equatorial Africa (James et al. 2014). Drying over west equatorial Africa can be large in some models (James et al. 2014), although the forest response is hard to predict.

The observations and field experiments summarised above have given greater confidence that forests are significantly affected by drought, emphasising the importance of extreme climate events in causing extensive tree losses (through drought and heat mortality and increased fire). On the other hand, some acclimation of trees to warming has been demonstrated. These vegetation responses are not well understood or represented in current dynamic vegetation models (e.g. Powell et al. 2013).

A recent study using three terrestrial biosphere models (Zhang et al. 2015) found the direction and severity of precipitation change to be critical. A greater model consensus for a projected lengthening and deepening of the dry season in Amazonia was found in CMIP5 compared with CMIP3 (Joetzjer et al. 2013), although it is unclear whether this represents a statistically significant improvement in model performance. A new observationally constrained model study (Boisier et al. 2015) found a greater lengthening of dry seasons over the Amazon than projected by unconstrained models (as found, with a different method, by Shiogama et al. 2011). A key uncertainty in terms of impacts is the extent to which forest whole-ecosystem responses to climate change might be protected by the wide functional diversity in many tropical forests. The work of Fauset et al. (2012) from Ghana suggests that by not accounting for biological diversity, most vegetation models may underestimate forest resilience.

In terms of land-use, it has become clearer that as well as direct deforestation, the indirect effects of deforestation on forest fragmentation, and on climate locally and downwind, must be considered in regulatory policies (e.g. Harper et al. 2014; Lawrence; Vandecar

2015). While a 70% decline has been reported in deforestation in the Brazilian Amazon between 2005 and 2013 (Nepstad et al. 2014), maintaining low levels of deforestation in a sustainable manner remains a challenge (Nepstad et al. 2014). Despite the reduction in deforestation since 2004, around half of the area burnt during 1999-2010 over the southern Amazon occurred during 2007 and 2010, when deforestation activity was relatively low, suggesting that fire-free land use needs to be encouraged as well as reducing direct deforestation (Morton et al. 2013). Achieving similar reductions in deforestation in other countries may be challenging due to issues with governance and monitoring capability (DeFries et al. 2013)

Various positive feedbacks (fire-vegetation and climate-vegetation; eg. Hirota et al. 2011; Hoffmann et al. 2012; Staver et al. 2011) exist that could lead to abrupt reductions in forest cover, for relatively small change in external forcings, and inhibit reversibility, but the processes are poorly characterised. The spatial scale over which abrupt or irreversible change might extend depends on the strength of these positive feedbacks, and demonstrating whether alternative stable states exist over large scales is challenging (Good et al. 2016). Indeed, recent observational work has challenged the notion that savanna and forest represent ‘alternative stable states’, since across the tropics multiple local gradations are almost always found (Veenendaal et al. 2015). Local fire-vegetation feedbacks are seen in prescribed burning experiments (Silverio et al. 2013), but over large scales, only 10% of the locations burnt in the 2005 drought showed repeated burning by 2010 (Morton et al. 2013). One new model study (Higgins; Scheiter 2012) found that, while transitions between vegetation states may be abrupt locally, over continental and larger scales the effect on the carbon cycle is much more gradual (because the timing of transitions varies with location). Another model study (Moncrieff et al. 2014) found that the area over which alternative stable states are possible could be large in present-day conditions, but declined substantially with future CO_2 increases. Hoffmann et al. (2012) noted that forest-fire feedbacks themselves can be sensitive to tree growth-rates – and hence to climate change.

d. Potential consequences

The observations summarised above give greater confidence that the Amazon represents a net carbon sink, but this appears to have been declining at least for a decade, and the long-term future of this sink is uncertain. Persistent drought would be likely to cause a transition to lower statured, lower biomass forest, from mortality of larger trees (Rowland et al. 2015). Extreme events could have a large impact on the global carbon cycle and offset or counteract potential increases in biomass (Reichstein et al. 2013).

While it is accepted that tropical deforestation tends to reduce evapotranspiration locally, consequent changes in rainfall are complex and depend on the scale and pattern of deforestation (Lawrence; Vandecar 2015). Including deforestation feedback on climate (via precipitation) is key in assessing river runoff change (Lima et al. 2014; Stickler et al. 2013). Stickler et al. (2013) estimate that when feedbacks on climate are included, the sign of change in hydropower generation potential for the plants under construction on Xingu River is reversed, declining to only 25% of maximum plant output under business-as-usual land-use projections by 2050. The net runoff response is basin-dependent (Lima et al. 2014) and is sensitive to the scale and pattern of deforestation (Lawrence; Vandecar 2015). Deforestation may reduce the length of the wet season, such that large-scale expansion of agriculture in Amazonia may be unsustainable (Arvor et al. 2014; Oliveira et al. 2013). Land-use-driven stream warming of at least 3-4K (in mean daily maximum temperature) in southeastern

Amazonian has also been observed (Macedo et al. 2013) - well above the ~1 K threshold for changes in fish physiology, growth and behaviour. Overall, multiple ecosystem services need to be taken into account when considering optimal management (Donoso et al. 2014).

e. Cautions (uncertainties)

Accurate projections are partly limited by the availability of observations. Inaccessibility of tropical forests increases reliance on remote sensing data, but also makes verifying remote sensing data (notably, precipitation, biomass, and vegetation productivity data) challenging. New studies have shown that great caution is required in interpreting satellite retrievals of variability in greenness (Morton et al. 2014). The tropical forest biome probably constitutes the largest terrestrial carbon sink, but it is also associated with the largest uncertainties (Pan et al. 2011), because of its great ecological complexity, huge scale, and multiple anthropogenic processes affecting it (Lewis et al. 2015).

There is substantial uncertainty in the CMIP5 projections of future precipitation in tropical forest regions, (Collins et al. 2013), although there is greater degree of inter-model agreement in some seasonal changes, such as a lengthening and a deepening of the dry season in Amazonia (Boisier et al. 2015; Joetzjer et al. 2013). However, the representation of present-day Amazon precipitation still contains large biases. Large uncertainties are also associated with the modelled response of vegetation to temperature (Galbraith et al. 2010; Huntingford et al. 2013) and to CO_2 (Rammig et al. 2010). Processes of direct mortality from fire and drought (and effects of fire on aerosol) are often either unrealistic or absent from models (e.g. Powell et al. 2013), and the range of plant functional types is extremely limited in relation to the large biodiversity and hence range of potential tree-level responses in most tropical forests.

f Comparison with AR5

The new literature has not altered the broad top-level view given in AR5. Probably the greatest advances lie in increased confidence that drought adversely affects the forest carbon balance – and improved understanding of how this occurs. Many uncertainties remain, and estimating the likelihood of basin-scale forest dieback remains challenging.

Ocean Acidification

a Introduction

Increased atmospheric CO_2 reduces seawater pH, increases the solubility of calcium carbonate (reducing saturation state), and causes other chemical changes together known as ocean acidification (OA). The biogeochemical, ecological and societal implications of OA have received greatly increased research attention during the past decade (Mathis et al. 2015; Riebesell; Gattuso 2015). OA risks and impacts were included as a component of climate change in the IPCC's Fourth Assessment Report, with more detailed analyses in the Fifth Assessment Report, particularly by Working Group II (IPCC 2014).

Analyses of geological OA events and modelling studies show that physico-chemical recovery from large-scale perturbations in ocean carbonate chemistry takes many thousands of years (Zeebe; Ridgwell 2011), due to slow rates of deep ocean mixing and of chemical equilibration with seafloor sediments. Thus longterm hysteresis effects are inherent in the response of global ocean chemistry to atmospheric CO_2 forcing, and there is only very limited capacity to accelerate future recovery by actively removing CO_2 from the atmosphere (Mathesius et al. 2015). Species' extinctions are necessary irreversible.

Many different thresholds or tipping points for OA impacts can be considered under conditions of steadily increasing atmospheric CO_2 levels; the focus here is on increased solubility of calcium carbonate (in particular, the saturation state for aragonite, the form of carbonate in the shells and structures of many marine organisms) and the risk of rapid loss of tropical corals.

b Observed recent changes

IPCC (2013) provided decadal measurements of ocean carbonate chemistry in near-surface waters at three oceanic monitoring sites; and other datasets are also now available (WMO 2014). All these observations unequivocally show decreasing pH in the upper ocean at rates (-0.0011 to -0.0024 yr^{-1}) closely matching those expected from rising atmospheric CO_2. Both physical and biological factors are responsible for the spatial and temporal variability in these datasets; whilst seasonality is usually smoothed-out for trend analyses (WMO 2014), it is of high ecological importance, determining the conditions experienced by marine organisms (Sasse et al. 2015).

There is much less temporal variability of pH in the ocean interior; however, there are also fewer longterm measurements. Atlantic observations (Woosley et al. 2016) confirm an anthropogenically-driven decrease in surface pH of ~0.0021 yr^{-1} with greatest changes in the top ~1000m; however, some decrease also occurs at greater depths. Such changes are superimposed on a natural decrease of pH with depth, with North Atlantic seafloor values generally being in the range 7.70 – 7.75 (Vazquez-Rodriguez et al. 2012).

Correlations between observed OA and biological or ecosystem changes are not necessarily causal, since other environmental factors are also likely to be involved. The strongest observational evidence relates to OA effects on pteropods (planktonic snails) in the Southern Ocean and northeast Pacific (Bednarsek et al. 2014b; Bednarsek et al. 2012); on cultivated oysters (Barton et al. 2015); on warm-water corals; and at natural CO_2 vents.

Longterm reductions of up to ~30% in the natural calcification and growth rates of tropical corals have been reported in several studies (e.g. Silverman et al. 2014). Linkage to OA has been demonstrated by *in situ* treatments of a natural coral community in the Great Barrier Reef (Albright et al. 2016). When water chemistry was restored to pre-industrial conditions by short-term alkalinity enrichment, coral growth rates increased by ~7%.

Observations at natural, shallow-water CO_2 vents consistently show marked decreases in overall biodiversity as pH declines (Fabricius et al. 2011; Hall-Spencer et al. 2008). Microbes in sediment are also affected (Raulf et al. 2015). Non-calcifying seaweeds and sea grasses out-compete calcifying organisms under such high CO_2, low pH, conditions, although some genetic adaptation of the latter can occur (Garilli et al. 2015).

c Potential for significant change

Experimental studies have shown that many marine species are likely to be negatively affected from future OA if high CO_2 emissions continue, with risk of ecosystem alterations at the global scale (CBD 2014; Gattuso et al. 2015; Nagelkerken; Connell 2015). Taxonomic variability in biotic responses to OA is, however, high; furthermore, complex interactions occur with temperature, food availability and other stressors (Kroeker et al. 2013; Ramajo et al. 2016; Wittmann; Portner 2013), and the potential for evolutionary adaptation is largely unknown (Sunday et al. 2014).

Marine ecosystems are susceptible to abrupt, non-linear changes (regime shifts; Mollmann et al. 2015) that cannot be easily reversed once thresholds, that may be of different kinds, are exceeded (Hughes et al. 2013; Mumby et al. 2011; Plaganyi et al. 2014). Two such OA-related thresholds were identified (Steinacher et al. 2013) in the context of allowable carbon emissions: aragonite undersaturation in the Southern Ocean, and the carbonate chemistry conditions necessary for warm-water coral reef survival.

Hauri et al. (2016) used a multi-model ensemble to determine changes in aragonite saturation state (Ω) around Antarctica and southern South America in an unabated CO_2 emissions scenario (RCP 8.5). The monthly occurrence of aragonite undersaturation (Ω <1.0) at the surface and at 100m water depth increased rapidly in most of these areas (Figure 4), particularly between 2040 - 2070 when atmospheric CO_2 levels are projected to be 500 - 650 ppm.

Closely similar effects are projected for the Arctic Ocean, where all surface waters north of 66˚ are projected to be unsaturated for aragonite by 2100 under RCP 8.5 (Popova et al. 2014). Regional differences are, however, greater - with surface undersaturation expected to have already occurred in the Siberian shelves and Canadian Arctic Archipelago (i.e. with current atmospheric CO_2 values of ~400 ppm), but not until the 2080s in the Barents and Norwegian seas (at ~ 900 ppm). The ecological significance of aragonite unsaturation is that such conditions are chemically corrosive to unprotected shells made of that form of carbonate, e.g. those of pteropods (Bednarsek et al. 2014a).

Coral exoskeletons are also made of aragonite: the depth distribution of coldwater corals is closely correlated with the aragonite saturation horizon (Guinotte et al. 2006; Jackson et al. 2014), whilst the calcification rate of both coldwater and tropical corals is sensitive to saturation state, responding semi-linearly over a wide range of values (Comeau et al. 2013;

McCulloch et al. 2012).

Most tropical coral reefs occur in waters where Ω > 3.0 (Manzello et al. 2014; Mongin et al. 2016), and that value has been used as a threshold for modelling climate change impacts (Steinacher et al. 2013). Whilst tropical coral growth can continue where Ω < 3.0 (Comeau et al. 2013; Shamberger et al. 2014), growth rates need to exceed bioerosion (Andersson; Gledhill 2013) and to be sufficiently rapid to allow reef recovery between temperature-induced bleaching events (Frieler et al. 2013). In theory, tropical corals could avoid the risk of bleaching by colonizing new sites where water temperatures have previously been too cool (Couce et al. 2013). However, the rate of current change may be too rapid for that to occur – and there are many geological precedents for 'coral reef crises', involving mass extinctions during geological warming and/or OA events (Kiessling; Simpson 2011). Based on these considerations, many coral researchers consider atmospheric levels of ~350 ppm CO_2 to be the 'safe' limit to ensure coral reef survival (ISRS 2015).

d Potential consequences

The potential consequences of OA are extremely wide-ranging, particularly for high emission scenarios. They include physico-chemical impacts (reduction in seawater capacity to absorb further CO_2); species-specific physiological and behavioural changes; perturbations in marine community processes, ecosystem functions and biogeochemical feedbacks; and changes in ocean ecosystem services, with societal effects on food security, coastal protection and climate regulation.

An overall reduction in marine diversity and abundances is expected to occur in a high CO_2 world (Nagelkerken; Connell 2015); nevertheless, not all species will be negatively affected. Some marine species that may be favoured also provide societal benefits, e.g. sea-grasses (Garrard; Beaumont 2014), but not all. Thus 'nuisance' species, such as jellyfish, seem generally tolerant of OA (Hall-Spencer; Allen 2015).

With regard to the carbonate undersaturation threshold identified above, the loss of pteropods from polar oceans would have wider consequences for food-webs, also affecting higher predators (fish, seabirds and sea mammals) of high commercial or conservation value, even if those groups are not directly affected by OA. Increasing OA in the Southern Ocean represents a risk to another key pelagic species, Antarctic krill. Krill eggs are sensitive to high CO_2 (Kawaguchi et al. 2013), and major reduction in their abundance would also jeopardise the entire ecosystem.

The potential loss of tropical coral reefs would have major consequences for coastal protection, tourism and fisheries, with the global economic value of those ecosystem services estimated to be up to ~ $1000 billion per year (Brander et al. 2012). However uncertainties in economic costs are high, and many other factors, in addition to OA, are affecting the future health and survival of coral reefs.

e Cautions (uncertainties).

Many uncertainties remain regarding OA impacts in the context of specific tipping points (Pandolfi 2015) and more widely (CBD 2014; Gattuso et al. 2015). The scaling-up of impacts from organisms to communities, food webs, ecosystems and economic impacts is challenging (Andersson et al. 2015; Ekstrom et al. 2015) – particularly since OA impacts do not act on their own, but co-occur with other stressors, both climate-related (warming, de-oxygenation and sea-level rise) (Gattuso et al. 2015; Howes et al. 2015) and non-climate-related (pollution, over-fishing and habitat loss) (Breitburg et al. 2015). Furthermore, coastal ecosystems seem likely to be at greatest risk from OA, but these are inherently complex and difficult to simulate in models because of interactions with sediment processes and riverine inputs (Artioli et al. 2014).

f Comparison with AR5

Since IPCC AR5, many OA studies have demonstrated variability in environmental conditions and biological responses, and the complexity of multi-stressor interactions. Such research therefore may seem to have increased, rather than reduced uncertainty. Nevertheless, understanding of OA and its impacts has significantly improved: observations have greater geographical coverage, integrating chemical and biological measurements, whilst new meta-analyses and assessments have confirmed previously-identified patterns and have also provided additional insights. Furthermore, greater attention has been given to important topics such as palaeo- OA events; socio-economic modelling; acclimatization and adaptation; and the vulnerability of cold-water corals.

Many of those more recent studies relate to the tipping points outlined here. In particular, there is now greater confidence that extensive aragonite undersaturation (with major ecological consequences) will occur in high latitudes if atmospheric CO_2 exceeds 450-500 ppm, and that warming will need to be well below 2˚C to avoid damaging interactions between ocean acidification and temperature for tropical coral reefs.

9. Conclusions

This report reviews the major new advances reported in the scientific literature, focussing on progress since AR5. The key findings are summarised in Table 1. Overall, compared to AR5, a large number of studies have added further detail to our understanding but the broad headline summaries of AR5 have not greatly changed.

For these systems, there is only limited quantitative information about the difference, in likelihood of significant change, between futures reaching 1.5 and 2K global-mean warming above pre-industrial levels. For ice-sheets and the effects of ocean acidification (combined with warming) on marine ecosystems, it is reasonable to assume that likelihood is higher for a 2K world than a 1.5K world. For Greenland, rates of mass loss and sea level rise are a non-linear function of the temperature increase (Applegate et al. 2015). A simplified model study of this ice sheet suggested that the global-mean warming threshold for irreversible loss has been estimated as only 0.8–3.2 °C (best estimate 1.6 °C) above pre-industrial (Robinson et al. 2012); while one long-term coupled model simulation found the threshold of zero surface mass balance crossed somewhere between 2 and 3°C above pre-industrial levels (Vizcaino et al. 2015). For ocean acidification, there is now greater confidence that extensive aragonite undersaturation (with major ecological consequences) will occur in high latitudes if atmospheric CO2 exceeds 450-500 ppm, and that warming will need to be well below 2˚C to avoid risk of damaging interactions between ocean acidification and temperature for tropical coral reefs.

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System	Key findings
Ice sheets	• From Greenland, the proportion of loss from surface melt has increased, becoming more consistent with long term model projections. The bedrock topography of the WAIS lends itself to an inherently unstable ice sheet. Some degree of irreversible loss may have begun, although the eventual magnitude and rate of this irreversible loss is uncertain. • There are indications that the East Antarctic Ice Sheet (EAIS) and the northeast Greenland ice stream may be more sensitive to climate change than previously expected. • New paleoclimate evidence for: 1) periods of relatively abrupt Antarctic mass loss following the last glacial maximum; 2) during the early Holocene (sustained warming ~2K above pre-industrial), WAIS mass loss rates comparable to present-day, but no WAIS collapse.
AMOC	• The observed AMOC overturning has decreased from 2004-2014, linked with decreases in subsurface density in the subpolar gyre. It is unclear at this stage whether this AMOC decrease is forced or is internal variability. • There was an unprecedented rise in US east coast sea level associated with the 2009-10 downturn in the AMOC.

Tropical forests	• Greater confidence that tropical forests are adversely affected by drought. • New climate models continue to suggest that basin-scale Amazon dieback from climate alone (as in the HadCM3 model) is not typical. However, these studies lack some key processes.
Ocean acidification	• Atmospherically-driven global trends in ocean acidification are superimposed on a dynamic natural system • Many factors affect variability in biological response; these are now much better understood • Extensive aragonite undersaturation in high latitudes can be expected if atmospheric CO2 exceeds 450-500 ppm, with major ecological consequences • Tropical coral reefs seem highly vulnerable to the interaction of ocean acidification and warming, with major economic consequences.

Table 1. Key new findings, for each system.

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2569-2587.

Figure captions

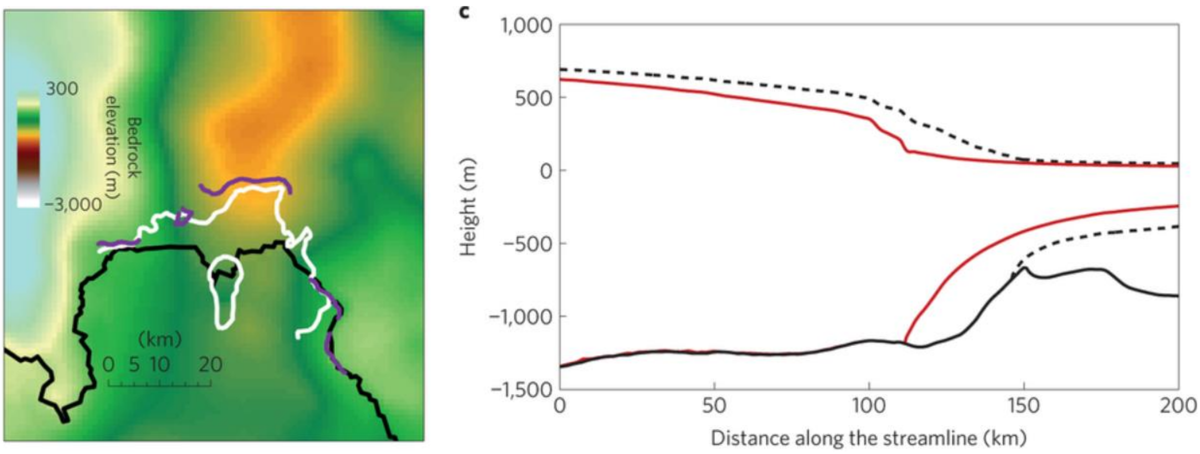
Figure 1. Evidence (from Favier et al., 2012) that the Pine Island Glacier's grounding line is probably engaged in an unstable 40 km retreat, due to the retrograde bedrock slope. Left: map of bedrock elevation, with the grounding lines for 2009 (white) and 2011 (purple) shown. Right: bedrock height (solid black line) and geometry of the glacier centreline produced by the Elmer/Ice ice-flow model at time (t) = 0 (dotted line) and after 50 years of a melting scenario (red line).

Figure 2. The 10 year time series of the AMOC measured at 26.5˚N (after Figure 2 from Srokosz & Bryden, 2015; original courtesy of David Smeed, NOC). The gray line represents the 10 day filtered measurements, while the red line is the 180 day filtered time series. Clearly visible are the low AMOC event in 2009-10 and the overall decrease in strength over the ten years.

Figure 3. Trends in net above-ground biomass change, productivity and mortality rates, for 321 plots, weighted by plot size (after Brienen et al., 2015).

Figure 4. Area-weighted ensemble mean duration (months per year) of aragonite undersaturation at the surface (solid lines) and at 100m depth (dashed lines) for three sectors of the Southern Ocean (Bellingshausen Sea, Weddell Sea and East Antarctica), the central Chilean coast and the Patagonian shelf. From Hauri et al (2016).

Figures



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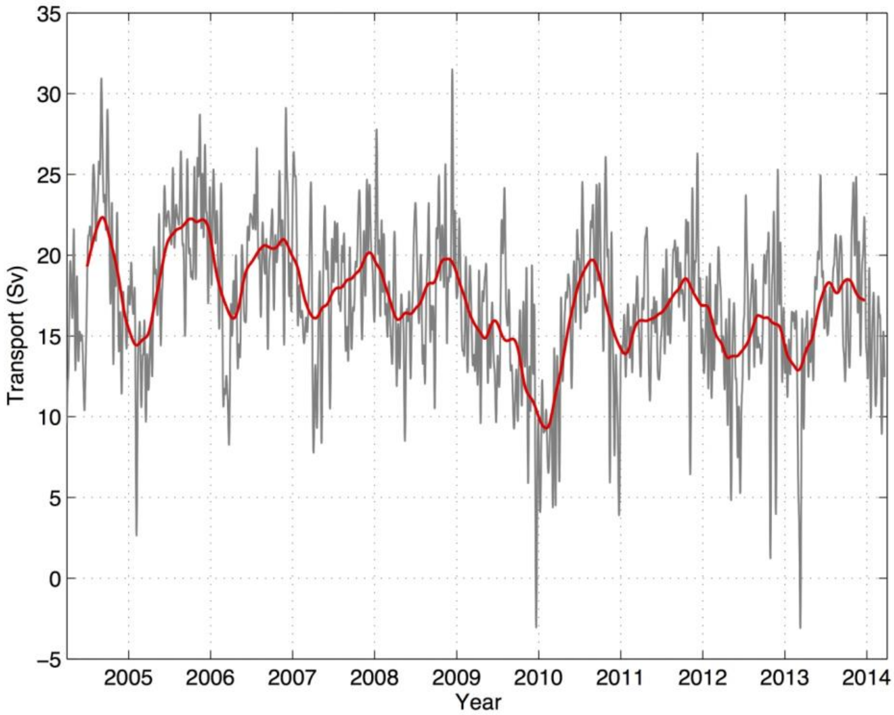

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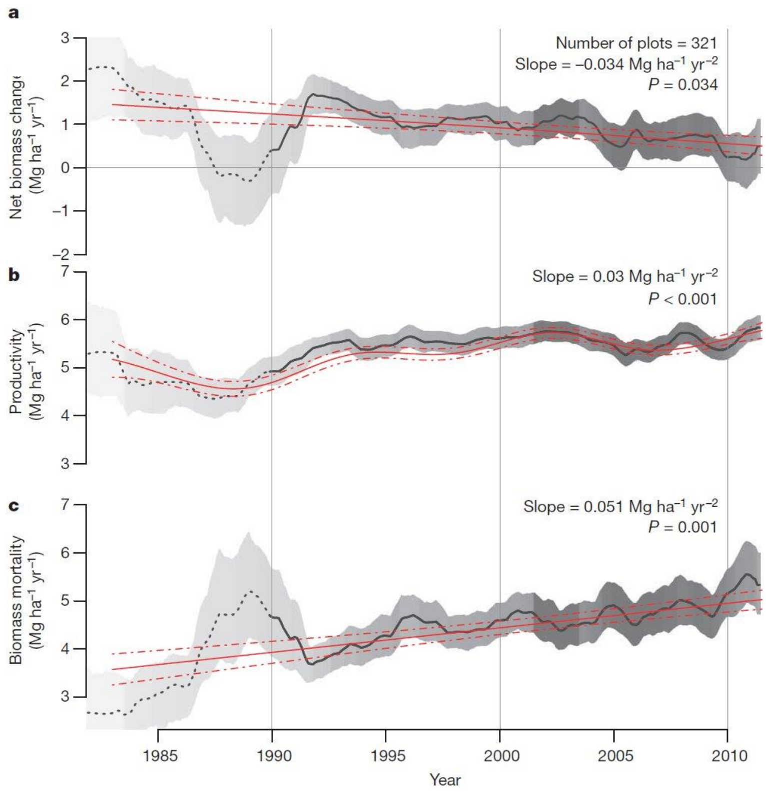

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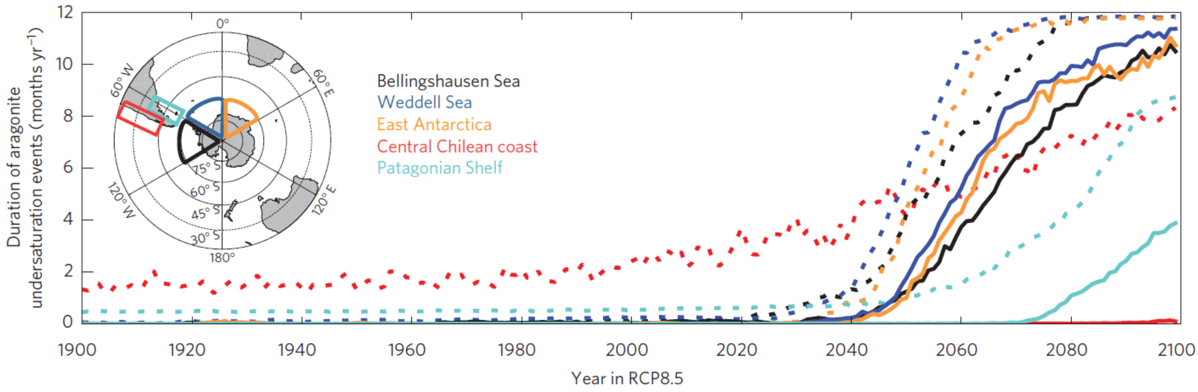

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